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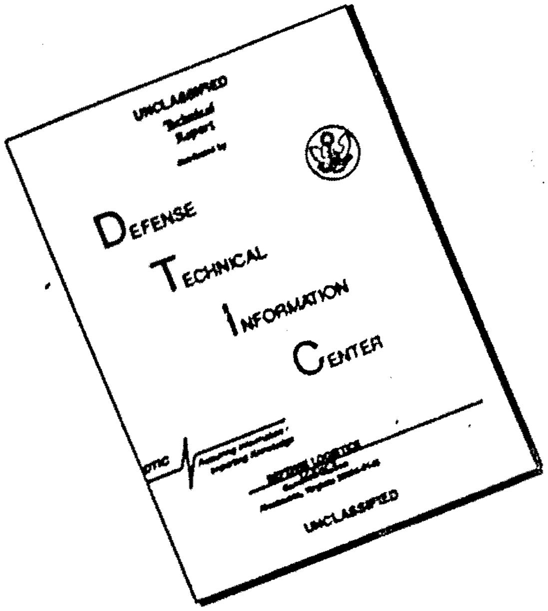

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Technical Report 2 April 1966

SEISMOMETERS ON ARCTIC ICE FOR DETECTION OF UNDERWATER ACOUSTIC DISTURBANCES (U)

Prepared for:

SUBMARINE ARCTIC WARFARE AND SCIENTIFIC PROGRAM
NAVAL ORDNANCE LABORATORY
WHITE OAK, MARYLAND

CONTRACT Nonr-2332(00)

STANFORD RESEARCH INSTITUTE

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By: NORMAN E. GOLDSTEIN

SRI Project ETU-2167-612

Approved: E. C. WOOD, MANAGER
GEOPHYSICS DEPARTMENT

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Abstract

In the Arctic Ocean the ice cover limits the usefulness of hydrophones because of the necessity to create or to locate openings in the ice through which to lower a hydrophone. On the other hand, the ice cover provides a convenient platform for seismometers, which are sensitive to the ice motions induced by an underwater acoustic source. Seismometers can be installed at almost any desired location on the Arctic Ocean, and if air dropped, numerous transducers can be installed rapidly. Furthermore, the Arctic Ocean provides an acoustical environment that permits the long range propagation of sound at frequencies for which the common exploration-type seismometer (geophone) is most sensitive, 10 to 100 cps. Because the wavelengths corresponding to these low frequencies are large compared to the average thickness of sea ice, and because the acoustic impedance mismatch between sea-water and sea-ice is small compared to the mismatch between sea-water and air, the ice should have little effect on the propagation of low-frequency acoustical energy through the Arctic Ocean. Thus, as a first approximation we can assume that the low-frequency particle velocities arising in sea ice from a distant underwater source should be the same as on the surface of an ice-free ocean with the acoustical properties of the Arctic Ocean.

Two environmental aspects seriously affect the value of seismometers on Arctic ice for detecting and locating an underwater acoustic source. First, within the 10 to 100 cps bandwidth, dynamic variations of 30 to 40 dB in the ice particle velocity (noise) have been observed. Estimates of signal amplitude have not yet been attempted, but these noise level variations suggest that, at times, unfavorable signal to noise ratios will be experienced. Second, as a result of the Arctic Ocean's velocity gradient, acoustical signals propagate dispersively, and this effect complicates the problem of locating the source.

Preface

This investigation was sponsored by the U.S. Naval Ordnance Laboratory, White Oak, Silver Spring, Maryland, under ONR Contract, Nonr 2332(00); SRI Project No. 2167-612. The study was conducted during the period 1 Nov. 1965 and 1 Mar. 1966.

The contract was monitored by M. M. Kleinerman, NOL. Project supervisor was W. H. Westphal and project leader was N. E. Goldstein.

Contents

Illustrations

Nomenclature

e	seismometer voltage output
h	damping coefficient of moving-coil seismometer
ω_N	resonant frequency of seismometer
y	displacement of seismometer case
G	intrinsic sensitivity of seismometer
φ_1	velocity potential
m	mode number
M	maximum mode of importance
$D_m(R)$	radial term of φ_1 for guided waves
$F_m(Z)$	vertical term of φ_1 for guided waves
R	range from source to receiver
ω	angular frequency
k	wave number
k_m	horizontal component of k, m-th mode
γ_m	vertical component of k, m-th mode
χ_m	angle of incidence of wave to surface, m-th mode
c_1	sonic velocity in the surface channel
Z_s	source depth
Z_r	receiver depth
$\vec{V}$	particle velocity
V_Z	vertical component of particle velocity
V_R	radial component of particle velocity
$W_n(Z)$	a function that depends on the physical parameter of the accustical environment and frequency

Nomenclature (Continued)

$\bar{\Phi}$	mean energy flux per cycle through a surface normal to the wave guide
$\bar{E}$	mean energy density between two vertical planes one guide wavelength apart
U	group velocity
N	number of seismometers in an array
θ	azimuth between source and array
ω_i	a particular angular frequency
t	time
Δt_c	time increment
ℓ	spacing between seismometers
n	n-th seismometer
c	phase velocity of an acoustic signal
C_{APP}	phase velocity of an acoustic signal in the direction of an in-line array of seismometers
S	sum of seismometer voltage amplitudes
A_n, B_n	voltage output from n-th seismometer
$c^{(m)}$	phase velocity, m-th mode
ψ	phase shift between modes at a range R from the source

Introduction

Passive detection of underwater acoustic disturbances has been a subject of intense study and has led to various hydrophone designs. The conventional hydrophone is sensitive to acoustical pressure variations, but there has been an attempt to use particle displacement hydrophones (Libermann and Rasmussen, 1964). In the Arctic Ocean the ice cover limits the usefulness of hydrophones because of the necessity to create or to locate openings in the ice through which to lower a hydrophone. On the other hand, the ice cover provides a convenient platform on which to locate passive transducers that are sensitive to ice motions. Seismometers, accelerometers, and strain gages can be employed for this purpose. Of these, the common exploration type of seismometer, often called a geophone, is particularly well-suited to the detection of an underwater acoustical disturbance becasue of its adequate sensitivity at low frequencies.

Basic Theory of a Seismometer

The most common exploration seismometer is the moving coil type (Fig. 1). These devices are small, rugged, and simple in principle. When coupled to an elastic medium the transducer's case will move in response to the particle motion of the medium. This motion causes an internal spring-supported mass and conducting coil to oscillate in the field of a magnet fixed to the case. If we neglect phase changes, hysteresis, eddy current, etc., the electromotive force (voltage) induced in the coil will be proportional to the time rate at which the coil cuts the lines of force. That is, the voltage output from the moving coil is

proportional to the velocity of the mass and coil. The expression for the voltage generated can be derived from the equation of motion of a damped linear oscillator and Faraday's law of induction:

$$\ddot{e} + 2h\omega_N \dot{e} + \omega_N^2 e = \dddot{y}\, G \tag{1}$$

where e = the voltage output

y = the case displacement

ω_N = the resonant frequency of the mechanical system = $2\pi f_N$

h = the damping term

G = the intrinsic sensitivity

The dots refer to differentiation with respect to time. G, the intrinsic sensitivity, is proportional to the number of turns on the coil and the magnet strength, among other things.

Two limiting cases of Eq. 1 are of interest. When the resonant frequency is very small compared to the excitation frequency (i.e., frequency at which the medium is oscillating) Eq. 1 reduces to

$$\ddot{e} = \dddot{y}\, G$$

For this condition the voltage output is proportional to particle velocity of the medium. This relationship is invalid when the resonant frequency is very large compared to the excitation frequency; then

$$e = \dddot{y}\, G/\omega_N^2$$

and the voltage output, although small, is proportional to the third derivative of particle displacement. At the frequency of maximum sensitivity, the resonant frequency, the voltage is proportional to a combination of particle velocity and acceleration.

Under all conditions if the electrical and mechanical characteristics of the detection system are known from calibration tests, it is possible to convert the observed voltages to displacements and velocities of the medium. This is often done in experimental and earthquake seismology. However, in the more applied field of seismology only the occurrence of a signal is of interest and a seismometer is selected to yield a maximum voltage output at the frequency of the expected ground motion signal.

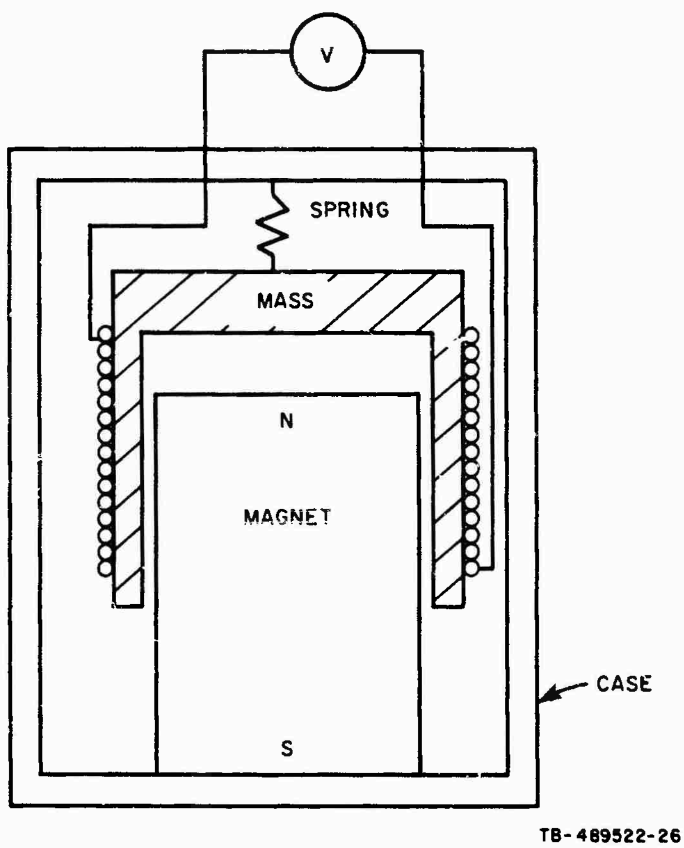

FIG. 1 COMMERCIAL MOVING COIL SEISMOMETER

Figure 2 shows sensitivity curves for a seismometer with a 28-cps resonant frequency and a 215-ohm coil resistance. Similar seismometers can be obtained with resonant frequencies ranging from 1 to 40 cps. With all these devices the sensitivity is sharply attenuated below the resonant frequency and the Q, or sharpness, of resonance can be varied by changing the load resistance, R, which in turn changes the total damping coefficient. The damping coefficients indicated in Fig. 2 are in percent of critical damping. For values of R near the coil resistance the device can be made to have a rather flat frequency response above the resonant frequency.

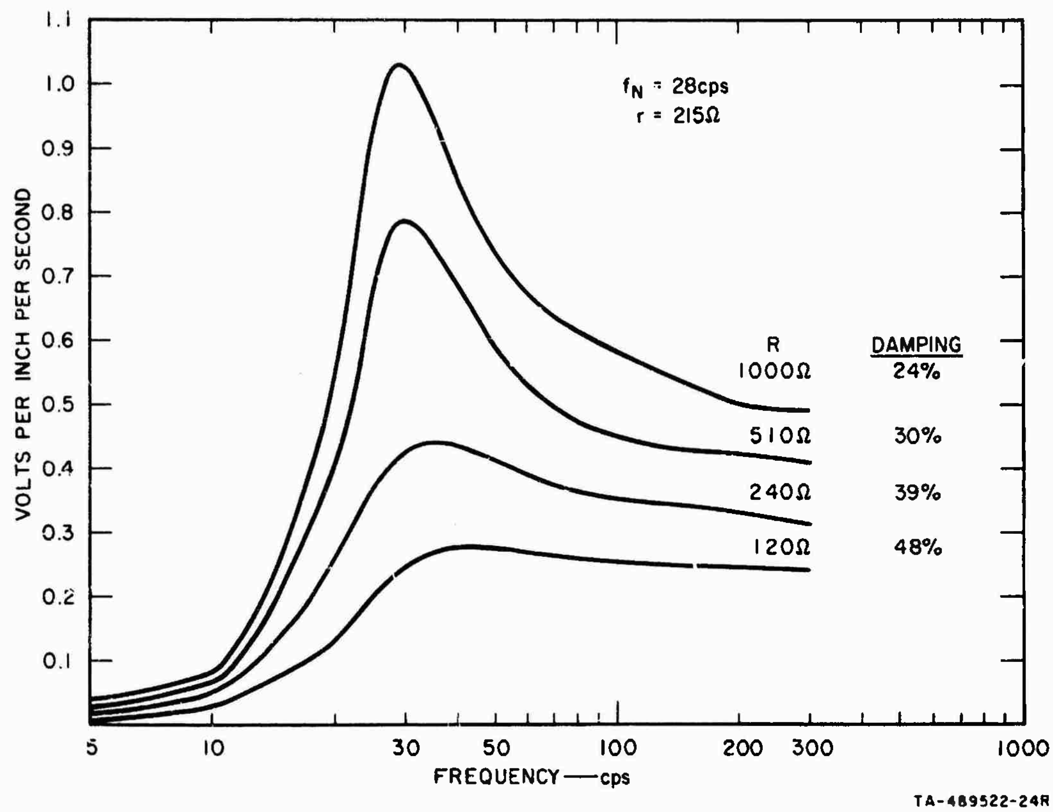

FIG. 2 SENSITIVITY CURVES FOR A HALL-SEARS SEISMOMETER

The common exploration seismometer was designed to be most sensitive to low-frequency excitation. This feature is of definite advantage if seismometers are to be used on Arctic ice because absorption of acoustic waves in sea water is negligible at low frequencies. Furthermore, low-frequency sound is not significantly scattered by ice keels and ice-bottom irregularities. The disadvantage of the moving-coil seismometer is that depending on its resonant frequency the signal output can be a complicated function of the first two derivatives of ice motion. In the following sections only particle velocities of Arctic ice will be considered.

Ambient Ice Noise

Arctic ice is dynamic: it is in continuous motion and there occurs within it an irregular background of particle motion. Clearly, the value of a seismometer to detect an underwater acoustic source is constrained by the amplitude of the ambient particle velocities, the noise level.

The frequency distribution of vertical ice displacements observed by Hunkins (1962) and Prentiss, et al (1965) are shown in Fig. 3. These data were calculated from observations made in April-May by Prentiss, et al and at various times of the year by Hunkins. In general the ice displacements decreased with increasing frequency of oscillation. At extremely low frequencies, 15 to 20 minutes in period, the displacements were of the order of centimeters. At frequencies above 1 cps the displacements approached the dimensions of gamma-ray wavelengths.

Figure 3 also shows attributed causes of the noise and the bandwidths affected. Below 10^{-2} cps the ice displacements have been attributed to atmospheric pressure variations. The displacements may be further enhanced by local wind action. Above 10^{-2} cps the noise spectrum is believed dependent on local wind (Prentiss, et al, 1965), but a definite correlation between noise and wind is not always clear.

At frequencies between 1 and 100 cps the noise is often transient with a non-Gaussian distribution of amplitudes. Prentiss, et al (1965) detected dispersive transient oscillations with frequencies of 1 to 10 cps,

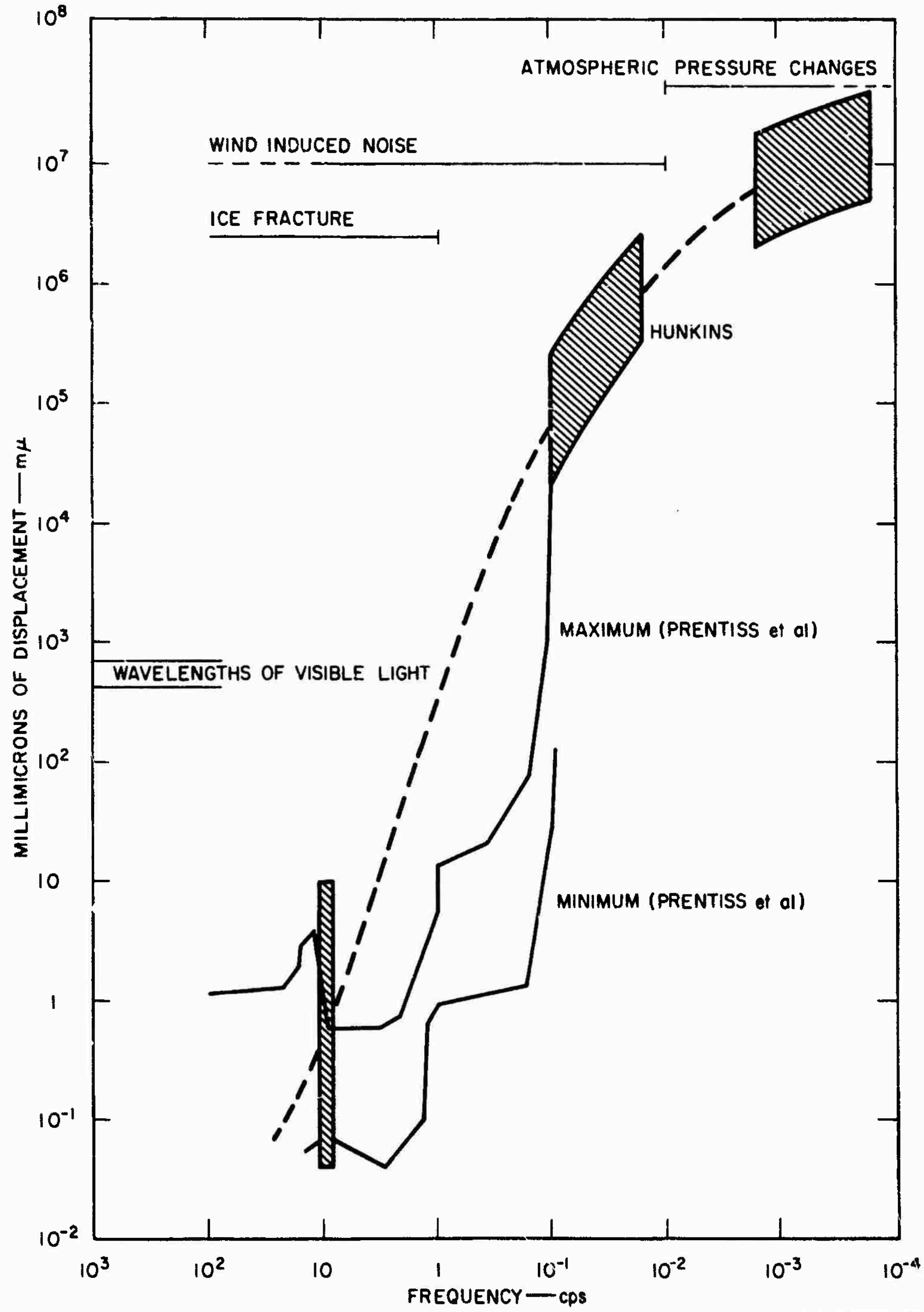

FIG. 3 AMBIENT ICE DISPLACEMENT IN THE ARCTIC OCEAN

which were thought to be from ice tremors. Milne and Ganton (1964) studied ice noise by means of bottom-mounted hydrophones beneath the Arctic ice cover and detected a significant amount of transient noise from the mechanical activity of the ice. The noise during late summer was attributed to the relative motions of the floes and during the winter to the cracking of the ice as a result of thermal stresses. These researchers also detected an irregular background of noise above 1 cps during the winter; this noise was tentatively attributed to wind-driven snow across the ice cover.

Because we may assume that the seismometer output is a voltage proportional to particle velocity, the displacement spectra were converted to velocity spectra (Fig. 4). The particle velocity noise also decreased with frequency, but the frequency dependence was not as pronounced as in the displacement spectra. Still, most of the noise was below 1 cps, and in the frequency range of the exploration seismometer (10 to 100 cps) the particle velocity noise was often no more severe than at quiet continental sites. There were periods, however, when the particle velocities above 10 cps were large in magnitude and might have interferred with the detection of a distant acoustic disturbance.

Our information concerning particle velocity of the noise is by no means complete. Yet, a sufficient effort has been made to study ice noise to emphasize the fact that we cannot easily predict its amplitude. There is evidence to suggest that a portion of the noise is caused by wind action and that the resulting noise spectrum depends on the properties of the coupled ice-to-sea-water-to-sea-bottom acoustical system. Thus, the noise spectrum will vary with water depth and acoustic impedance of the sea-bottom sediments. There is also evidence that the particle velocity of the noise in the bandwidth of interest (10 to 100 cps) has a diurnal variation during the winter: the noise increases toward evening when the air temperature drops and resulting tensile stresses cause cracking of the surface. Furthermore, Milne and Ganton (1964) have stated that there is a seasonal variation of particle velocity noise, the average noise level being larger in the winter than in the late summer.

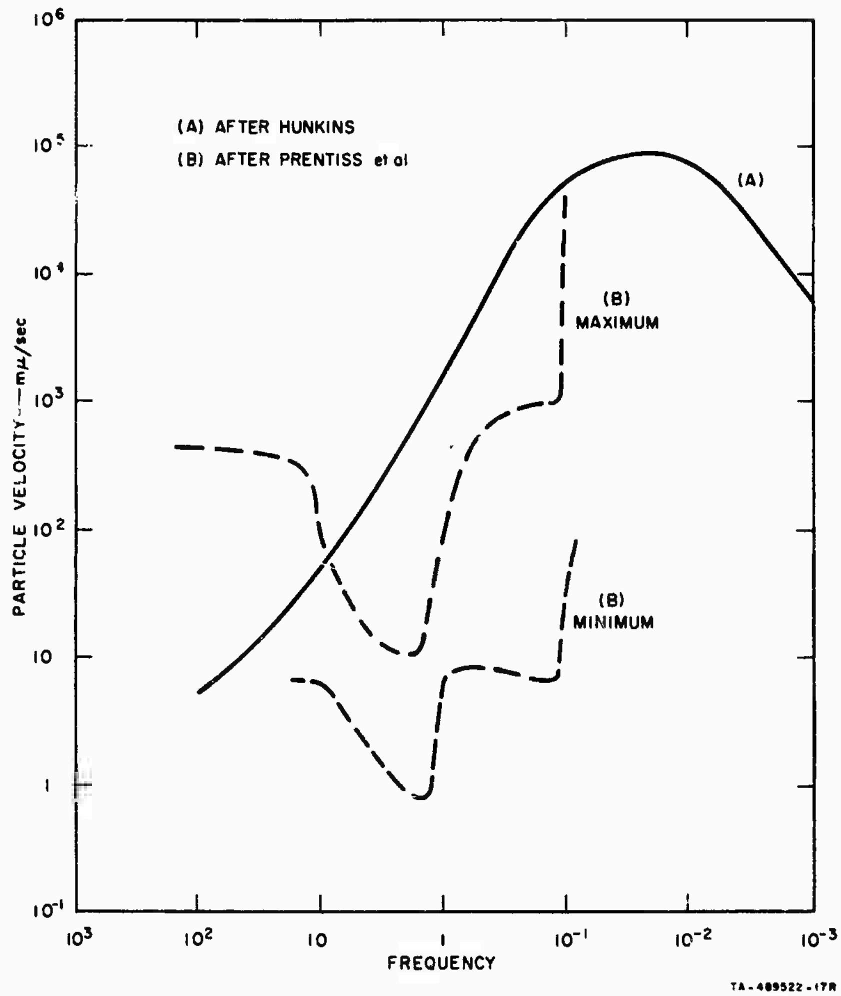

FIG. 4 AMBIENT VERTICAL PARTICLE VELOCITY OF ARCTIC ICE

Propagation of Particle Velocities in the Arctic Ocean

In regard to hydrophone operation a considerable effort has gone into determining the depth beneath the water surface at which the signal-to-noise ratio might be maximized. A seismometer, however, is restricted to operation either on the ocean bottom or on the ice cover above. Fortunately, the ice cover provides a satisfactory location to observe particle motion induced by an underwater source in the Arctic Ocean. This fact will be developed from a consideration of the Arctic Ocean acoustical environment.

The Arctic Ocean acoustical environment is inhomogeneous and shows vertical variations of sonic velocity and density of the ocean and sub-oceanic rocks (Fig. 5). For the ocean these data are based on vertical temperature and salinity profiles and were given by Kutschale (1961). For the sub-oceanic rocks these values are based on seismic refraction studies off the coast of Alaska (Schor, 1962). Although the sub-oceanic data were obtained from studies in the Bering Sea, our supposition is that data from beneath the Arctic Ocean would not be drastically different. Our knowledge of the acoustical environment beneath the Arctic Ocean will be improved upon the completion of the analysis of recent refraction surveys made on Arctic ice (Ostenso, personal communication).

The major feature of the acoustical system shown in Fig. 5 is that directly below the surface there exists a low-impedance channel bounded above by air and below by higher impedance media. Let us ignore the ice layer for the present. Although the lower boundary of the channel is not sharply defined, the low-impedance surface channel behaves as a wave guide. Thus, acoustical energy emitted within the channel will propagate along particular ray paths, on which constructive phase interference takes place upon reflection of the rays from the free surface and from the higher impedance medium at depth. For a given frequency there exists a multiplicity of ray paths that satisfy the condition for constructive interference. Each path corresponds to a particular mode (normal mode) of propagation. This kind of propagation is dispersive, i.e., the wave guide velocity (phase velocity) depends on frequency and the angle of incidence between the ray path and the reflecting surfaces.

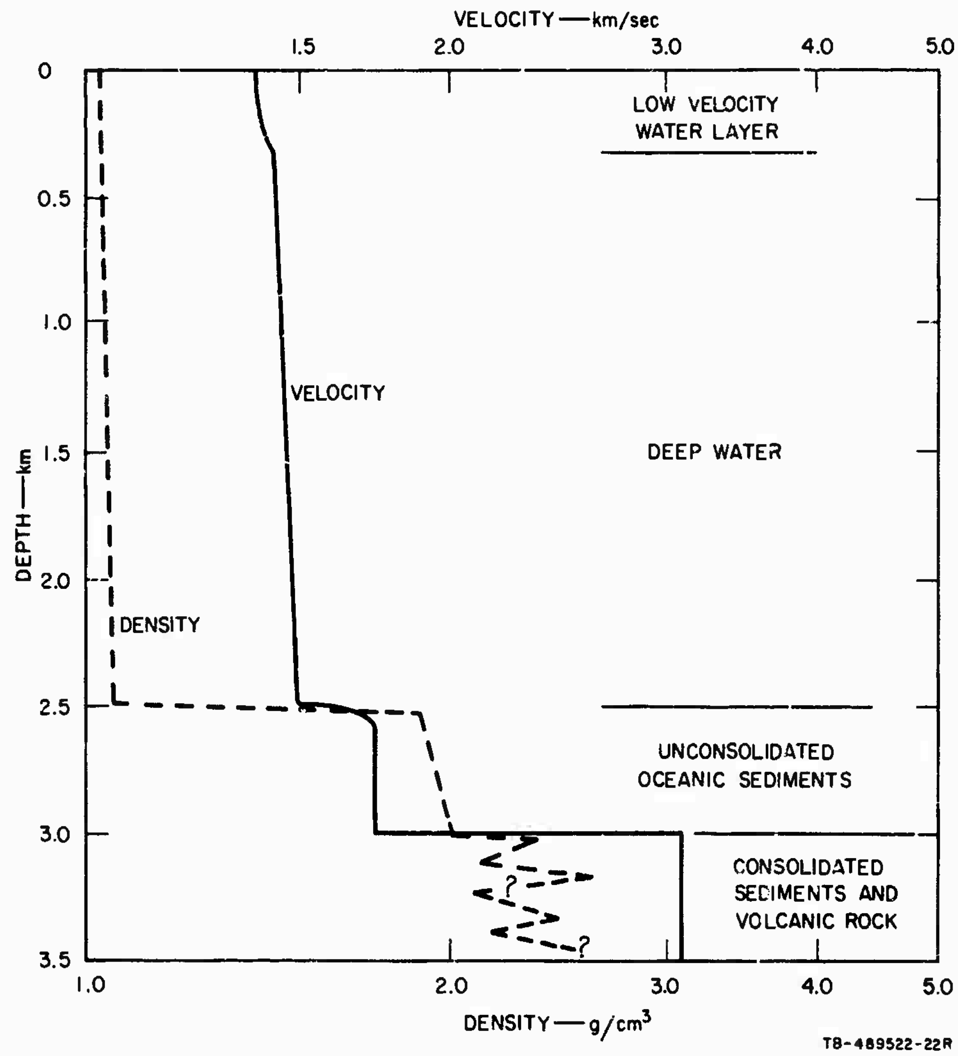

FIG. 5 VELOCITY AND DENSITY VARIATION IN THE ARCTIC OCEAN

In seeking a solution for the particle velocities in the wave guide we may begin by assuming a point source acoustical radiator located within the low-velocity surface channel. If the inhomogeneous ocean is divided into N horizontal layers, each with a constant velocity and density, and if we consider only the contribution from guided waves, i.e., waves unattenuated by incomplete reflections at any interface, then the velocity

potential in the near-surface channel is given by Tolstoy (1955, 1958) as

$$\varphi_1 = - \sum_{m=1}^{M} D_m(R)\, F_m(Z)\, e^{i\omega t} \tag{2}$$

$D_m(R)$ is a term dependent on the horizontal range, R , between source and receiver

$$D_m(R) = R^{-1/2} \exp\left(-i\left(k_m R + \pi/4\right)\right)$$

where k_m is the horizontal component of the wave number, k , in the surface channel that leads to the m-th normal mode of propagation. Expressed in terms of angle of incidence

$$k_m = \frac{\omega}{\alpha_1} \sin \chi_m$$

where ω = 2 πf

α_1 = sonic velocity in the surface channel

χ_m = the angle of incidence of the ray for which normal mode propagation is possible

The term $F_m(Z)$ in Eq. 2 depends on the source and receiver depth and the physical dimensions and elastic parameters of the layers. Its exact analytical expression is somewhat complicated but it is of the form

$$F_m(Z) = \frac{W_m(Z)}{\omega} \sin(r_m Z_s) \sin(r_m Z_r)$$

where Z_s = source depth

Z_r = receiver depth

$r_m = (k^2 - k_m^2)^{1/2}$

From Eq. 2 the particle velocities in the vertical and radial directions are

$$\vec{v} = \vec{\nabla} \psi,$$

$$V_Z = - \sum_{m=1}^{M} r_m \cot(r_m Z_r) \, D_m(R) \, F_m(Z) \, e^{i\omega t} \tag{3a}$$

$$V_R = - \sum_{m=1}^{M} F_m(Z) \, D_m(R) \cdot \left\{ \frac{1}{2R^{1/2}} - i k_m \right\} e^{i\omega t} \tag{3b}$$

Equations 3a and 3b are sufficiently general provided R is at least several times the low-velocity channel thickness and M is large, particularly at small R where the modal solution must approach the ray solution for which an infinite number of ray paths exist.

Although the numerical evaluation of Eq. 3a and 3b are best accomplished with a high-speed computer, useful qualitative information can be derived from the expressions. For example, as the point of observation approaches the surface, V_R approaches zero but V_Z does not. Moreover, for given Z_s and R the modulus of the complex function V_Z is a maximum at the surface. Another feature of interest is that the attenuation due to spreading for V_Z has only a $R^{1/2}$ dependence; the waves spread cylindrically. Lastly, the particle velocity V_Z is frequency dependent for a source whose power output is constant at all frequencies. This is due to the term W_m which, as a function of the physical parameters of the acoustical wave guide, indicates how well energy at any frequency is permanently trapped in the low-velocity surface channel by continuing total reflection.

If we know explicitly how W_m varies with frequency we can choose a transducer with a maximum sensitivity in the frequency range for which W_m is also a maximum. This is not a particularly straightforward task, but we can examine the form of W_m qualitatively.

A general theorem given by Biot (1957) states that the velocity of energy transport is equal to the mean energy flux per cycle through a surface normal to the wave guide ($\bar{\Phi}$) divided by the mean energy density between two such planes one guide wavelength apart ($\bar{E}$). If our source is a harmonic oscillator, then the velocity of energy transport is the group velocity U. Thus

$$U = \frac{\bar{\Phi}}{\bar{E}} \tag{4}$$

From this expression it can be seen that maximum energy densities will occur at minima of the group velocity. For the Arctic Ocean environment (Fig. 5) the group velocities of the first four normal modes are shown in Fig. 6 (Kutschale, 1961). In this environment the group velocities decrease with increasing frequency. If the calculations had been carried out for frequencies higher than considered in Fig. 6 then it could be observed that the velocities oscillate somewhat and approach the velocity of the surface channel at a sufficiently high frequency. At long ranges from the source only the lowest modes are important because the higher modes correspond to rays that undergo many more reflections and suffer a greater attenuation. As a first approximation we can therefore assume that the energy transported in the first mode will contribute most to the surface particle velocities.

There is a suggestion that for the first mode a group velocity minimum may occur somewhere in the 30 to 50 cps bandwidth. This is a reasonable assumption for we know that U cannot decrease monotonically with frequency; it must approach asymptotically the velocity of the upper layer which in this case is 1440 m/sec. Thus on the basis of Eq. 3 we can conclude that a maximum particle velocity should be observed above 30 cps. Experimental evidence (Kutschale, 1961; Greene, 1965) supports this conclusion.

Therefore, if we seek the particle velocity signal from distant underwater source whose output spectrum extends above 30 cps, we might employ a vertical-component seismometer with a frequency-dependent

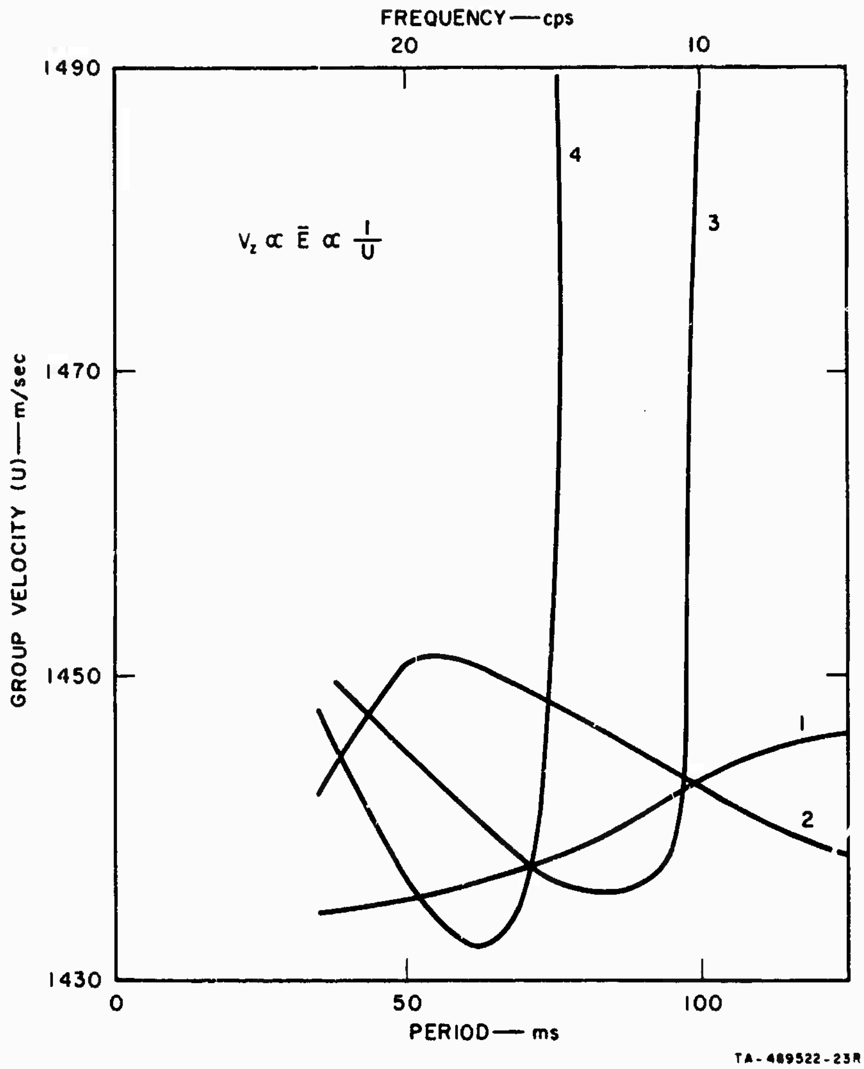

FIG. 6 GROUP-VELOCITY DISPERSION FOR THE FIRST FOUR NORMAL MODES IN THE ARCTIC OCEAN (after Kutschale)

sensitivity as shown in Fig. 2. Furthermore, a distinction between signal and noise might be achieved on the basis of particle orbits because the signal we seek has only a vertical component at the surface whereas the noise from sources closer to us will have both horizontal and vertical components.

Effect of Arctic Ice on Low-Frequency Particle Velocities

Thus far we have ignored the effect of ice on the particle velocities induced by a distant underwater source. In the Arctic Ocean the ice cover does modify the acoustical system, and we cannot entirely dismiss the effect of the ice and its corrugated bottom on the propagation of particle velocities. At long distances from the source, Eq. 2 is only approximate because the ice-bottom irregularities scatter the incident acoustic waves, thereby increasing the attenuation. However, at low frequencies, where the wavelengths are much greater than the ice thickness or the ice-bottom irregularities, the ice should only slightly modify the pressure and velocity fields in the ocean. To illustrate this, the reflection coefficients for a plane acoustic wave incident from sea water onto a flat ice layer 3 meters thick were calculated (Fig. 7). Incident angles of 75, 80 and 85 degrees were taken because this angular range corresponds to ray-path directions of the lower modes. As the reflection coefficients are no longer -1, the introduction of the thin ice layer changes the free surface boundary condition. However, at low frequencies, at or below perhaps 30 cps, the reflection coefficient is sufficiently close to -1, and for these frequencies the free surface is a reasonable approximation. Thus, we might suppose that the low-frequency vertical particle velocities at the ice surface should be little affected by the existence of the ice.

There is some experimental evidence, however, to indicate that the surface particle velocities at 30 cps are modified by the ice. Green (1965) found that the vertical particle velocities from a distant (330 nmi) underwater explosion were 6 dB greater on ice 0.1 meter thick than on nearby ice 4 meters thick. The marked difference was attributed to attenuation by the ice. If this is true the material has a remarkably high attenuation value of 1 dB/m. Absorption of sound in sea ice is apt to be high and variable because of the fractures and pores containing air and liquid brine, but the amplitude difference observed seems too large to be explained by absorption alone because the ice thicknesses involved are only a small fraction of a wavelength. Presently we have little experimental evidence upon which to derive a quantitative relationship between

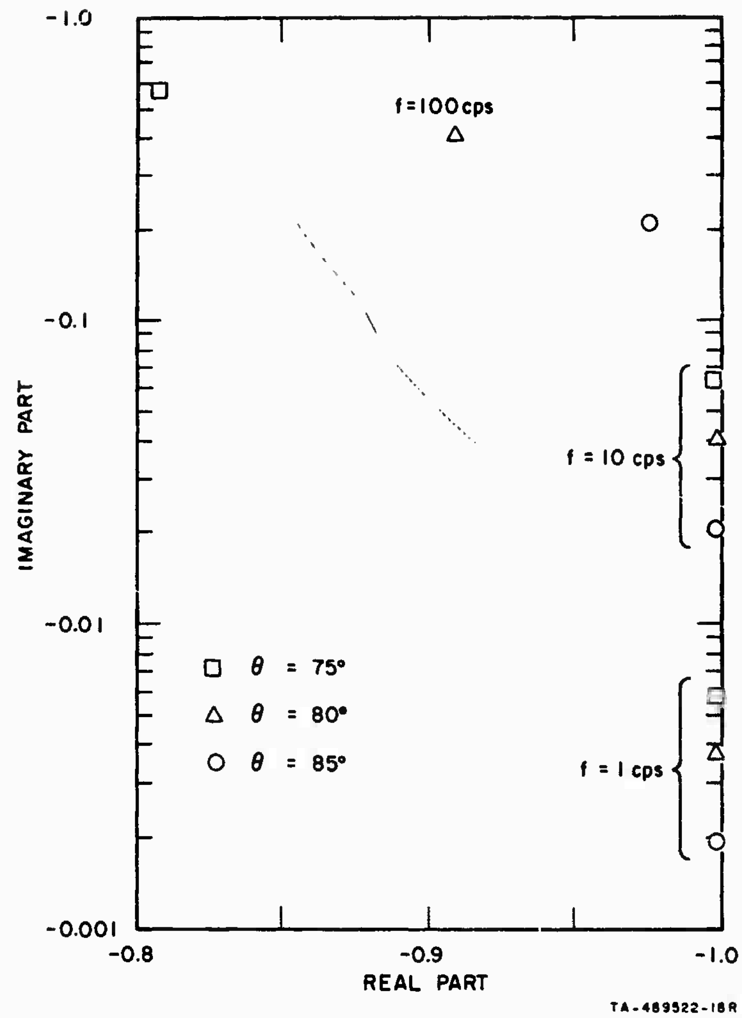

FIG. 7 REFLECTION COEFFICIENTS FOR A SOUND WAVE INCIDENT FROM SEA WATER ON ICE 3 METERS THICK

ice thickness and signal amplitude. From theory alone the low-frequency particle velocity amplitudes should be little affected by the ice unless the ice is at least several times the average Arctic ice thickness.

Seismometers Air-Dropped onto Arctic Ice

The use of seismometers on Arctic ice for surveillance would achieve maximum effectiveness if they, together with the associated electronics, could be air dropped. Air dropping the electronics should pose no special problem, but for the seismometer to be useful it must be both well-coupled to the ice and properly oriented after impact. The orientation is important because the seismometer will operate either at a reduced sensitivity or not at all if the mechanical axis deviates by more than about 10 degrees from the vertical.

Consideration of the problems of air dropping has prompted us to conceive a number of simple systems that might overcome these difficulties. Shown in Fig. 8 is a schematic representation of one such system. In this example, gravity orients the seismometer after impact. Within the enclosure containing the seismometer there is a low temperature alloy, such as Wood's metal, or water premelted by the heating coils before the air drop. Upon cooling, the alloy or water freezes the seismometer in place; thus seismometer orientation is fixed and coupling is made to the enclosure.

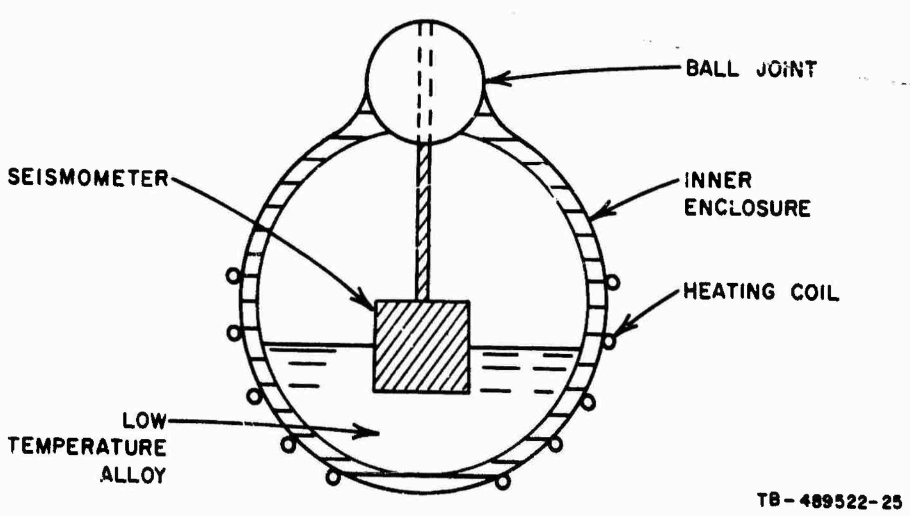

FIG. 8 .HEMATIC OF A SEISMOMETER HOUSING FOR AIR-DROP OPERATIONS

Not shown in the figure is an outer casing that might have an aerodynamic shape, guiding fins, and a nose spike to penetrate the surface layers of ice and snow. If a low temperature alloy instead of water is used the temperature of the outer casing would be significantly increased due to the heat supplied within it. Although some of this heat would be radiated into the air, it would also melt the surrounding ice. Upon refreezing, the ice would be in good mechanical coupling to the seismometer case. An engineering problem is to design the system so that the ice surrounding the outer casing freezes before the seismometer is frozen in place.

Use of Geophone Arrays

The problem of locating an underwater, continuous acoustic source by means of seismometers poses difficulties that are not encountered in other fields of applied seismology. The continuity of the signal precludes the use of first-arrival information that seismologists often rely on for locating earthquakes and man-made seismic disturbances. Moreover, particle velocity amplitude information has marginal value as a guide to the source location. The near-surface, low-velocity channel of the Arctic Ocean creates a phase velocity dispersion of the acoustic signals, and, as a consequence, particle velocity intensities at frequencies of 30 to 100 cps oscillate appreciably with range (Pederson and Gordon, 1965, Tolstoy, 1966). Unless we were to cover a large region with a closely spaced network of detectors we could not hope to determine the source location on the basis of seismometer voltage amplitudes alone.

With a limited number of seismometers the only certain means for locating the source is an analysis of the phase characteristics of the particle velocity signals from an array of seismometers. This will not be a trivial processing procedure. As an example of one such array we can think of a group of N seismometers arranged in a semicircle with an additional seismometer, the reference seismometer, at the center of the equivalent circle. If a continuous acoustic source is at a moderately long range from the array, then the surfaces of equal phase will be

planar, and these planes will continuously sweep past the array. The source direction can be obtained by comparing the phase relationships of seismometer outputs. That is, the seismometer whose output has the greatest coherence with the reference seismometer output lies on a radius of the semicircle that is parallel to the planes of equal phase. Hence, if a number of such arrays are used the source position can be estimated by a simple triangulation of the lines normal to the radii of equal phase. The physical limitation of this approach is that if we have no prior knowledge of the source position we are obligated to use a very large number of seismometers; otherwise our resolution of the source direction, which varies as π/N, will be poor. If an azimuth resolution of 5^0 is desired, 36 seismometers would be necessary.

Good resolution with a small number of seismometers requires an array that is steerable. To illustrate the steerable array and its limitations we consider a method by which the angle, θ, between a line of N detectors and the source can be determined. We again assume a continuous source at a moderately large distance from the array so that the surfaces of equal phase are planar. If the geophone outputs are narrow-band filtered about a frequency ω_0 and then summed, the resulting signal will be

$$S = \sum_{n=0}^{N-1} A_n \cos\left(\omega_0 t \pm \frac{n\omega_0 \ell}{c_{APP}}\right) \tag{5}$$

where ℓ is the distance between geophones and c_{APP} is an apparent phase velocity, the phase velocity of the wave fronts as seen in the direction of the linear array. This velocity is related to the true phase velocity, c, by the expression $c = c_{APP}(\cos\theta)$. Provided the true phase velocity at frequency ω_0 is known, the angle θ can be found by means of delay filtering the geophone signals prior to the summation. That is, if we delay each signal by an amount $\pm n\Delta t_0 (n = 0$ to $N-1)$, we should find a particular Δt_0 for which all the geophone outputs are in phase and their sum is therefore a maximum. The particular Δt_0 that results in a

maximization of S is simply related to θ by

$$\pm \Delta t_0 = \frac{\ell \cos \theta}{C_{APP}} \tag{6}$$

The plus or minus sign is carried to indicate that the direction of increasing delay will depend on whether seismometer 1 or seismometer N is closer to the source. Equation 6 is independent of N; however, larger N's would help to reduce random noise effects that would otherwise interfere with the determination of the Δt_0 we seek.

To determine whether θ is to be measured clockwise or counterclockwise from the line, we require additional information which can be obtained by means of other arrays.

The theoretical difficulty of utilizing the technique just described is that more than one mode can exist at frequency ω_0, and each mode of propagation will have associated with it a different phase velocity. For example, if the particle velocity is propagated in only two modes, then Eq. 5 must be written as

$$S = \sum_{n=0}^{N-1} A_n \cos(\omega_0 t \pm \nu_n) + B_n \cos(\omega_0 t \pm \mu_n) \tag{7}$$

where A_n and B_n are the amplitude of the particle velocity in each of the modes and

$$\nu_n = \frac{n \omega_0 \ell}{C^{(1)}_{APP}}$$

$$\mu_n = \frac{n \omega_0 \ell}{C^{(2)}_{APP}} + \varphi$$

φ = phase shift between the modes.

Thus, a Δt_0 might be found that maximizes Eq. 6 but it will no longer be simply related to the angle θ because φ, the phase shift between modes, is dependent on the distance between the source and the array. If we examine just the radial term, D_m, of the particle velocity (Eq. 2) it can be shown that the phase difference between the first two modes is

$$\varphi = \omega_0 R \frac{c^{(2)} - c^{(1)}}{c^{(2)} c^{(1)}} \tag{8}$$

where $c^{(1)}$ and $c^{(2)}$ are the phase velocities of each mode and R is the radial distance from the source.

Qualitatively, there exist two possible solutions to the multimode difficulty as it affects the determination of θ. One solution is to select a particular frequency, ω_0, such that the excitation function for the first mode is very much larger than these functions for all other modes. The second solution is to select a particular frequency such that the phase velocities in all modes are equal. Of these two approaches only the first might be successfully employed, but a calculation of the excitation functions for the Arctic Ocean environment would be necessary to determine whether the approach is feasible. The second approach is unworkable from a quantitative standpoint because calculations made by Kutschale (1961) for the Arctic Ocean environment indicate that the deviation between phase velocities for all modes becomes small at frequencies well above a few hundred cps.

Conclusions

Particle motion transducers embedded on Arctic ice to detect man-made underwater disturbances overcome a disadvantage of Arctic hydrophone operation, which requires the lowering of hydrophones through a hole in the ice cover. Seismometers can be installed at almost any desired location on the Arctic Ocean, and, if air dropped, many transducers can

be installed in a short time. Furthermore, the Arctic Ocean provides an acoustical system that permits the long range propagation of vertical particle velocities at frequencies for which the common exploration seismometer (geophone) is most sensitive, 10 to 100 cps. In this bandwidth the ambient ice noise is variable in amplitude. Experimental evidence shows periods of quiescence occur during which the particle velocity noise is no more severe than at quiet continental sites. A few hours later, however, the ambient noise has been observed to increase by a whole order of magnitude over the quiescent level. Although limited, our present knowledge of the particle velocity noise amplitudes suggests that the detectability of a submarine will be greatly dependent on the local noise conditions. Just how unfavorable the signal-to-noise ration will be under any given set of noise conditions is not known at this time. However, a quantitative determination of the signal-to-noise ratio in the Arctic environment will be one objective of the continued analytical study.

To acquire quantitative information an *in situ* experiment would be the preferred approach, but such an experiment in the near future is unlikely. Hence, anticipated signal-to-noise ratio can only be surmised from observed noise levels and calculated signal levels. A computer program is now being written to provide us with the theoretical signal (particle velocity) amplitudes from an underwater point radiating an arbitrary acoustic power spectrum. Because the properties of the propagation medium, the Arctic Ocean, and the characteristics of the noise are variable in time and space, the calculated signal-to-noise ratios can never be uniquely defined through analytic techniques. Nevertheless, the calculated signal-to-noise ratios will be helpful for quantitative identification of specific problems related to the usefulness of seismometers on Arctic ice.

The ice layer presents a subject for concern not only because it is a source of noise, but because under some conditions it affects the propagation of particle motion in the Arctic Ocean. However, it was found than when the ice was less than 3 meters thick an acoustic wave whose

frequency is less than 30 cps would be almost totally reflected, suffering only a 180^{0} phase change. Thus, under these ice conditions the ice layer exerts a negligible influence on the incident wave, and the ocean surface may be analytically treated as a free surface. This conclusion is substantiated by Kutschale (1961) who found the observed phase velocities of low-frequency acoustic waves in the Arctic Ocean to be insignificantly different from the velocities calculated with the free surface approximation. At frequencies less than 30 cps, then, the particle motion at the surface of the ocean should be modified little by a thin ice layer, and there should be, effectively, a good coupling of particle motion to the ice.

The study to date has not considered the coupling of acoustic wave energy to the ice at frequencies greater than 100 cps. For high-frequency waves, whose wavelengths in sea water approach the dimensions of the average ice thickness and the ice-bottom irregularities, we cannot ignore the influence of the ice on the phase velocities and amplitudes of guided wave propagation. In addition to the effect of the ice on their propagation, higher frequency waves are sensitive to small spatial and temporal variations in the oceanic density and compressional wave velocity, and hence the particle motion of these waves cannot be accurately predicted from a numerical analysis.

The problem of locating an underwater sound source with seismometers is basically no different from the location problem involving hydrophones. A method is required that is independent of the source output power and relatively insensitive to variations in the properties of the propagation medium. Moreover, the method must be capable of resolving the source direction to within a small percent of 2π radians. In this regard, the characteristics of the Arctic Ocean work to both our advantage and disadvantage. The roughly bi-linear, positive velocity gradient in the ocean causes the particle velocity radiated from a near-surface acoustic source to attenuate approximately as the inverse of distance to the one-half power. On the other hand, the velocity gradient produces a dispersion of the radiated signal. As a consequence of this dispersion, a single frequency propagates at a finite number of phase velocities; each phase

velocity corresponds to a normal mode of propagation. Because the phase relationships between the modes are functions of range, the signal intensity is a complicated, multiple-valued function of range due to the phase interference of the multiple modes. The complicated nature of the signal makes it difficult to predict how best to process the outputs of seismometers in arrays as a means for determining the direction to the source. One technique that offers the possibility of good resolution of the source azimuth is the steerable array of seismometers. This technique will not yield interpretable results, however, unless there is a frequency, within the frequency band of maximum acoustical radiation by the source, at which the particle velocity signal is almost entirely confined to a single mode. The computer program under development will serve to answer the question of whether steerable arrays might be used.

(C) Two areas of Arctic experimental work have not received adequate attention. First, no thorough study of particle velocity noise in the frequency range of 1 to 100 cps has been conducted. We have little knowledge concerning the orbital motion of the ambient noise and the diurnal and seasonal variation of noise intensities. Moreover, no attempts, to our knowledge, have been made to experiment with arrays of parallel connected seismometers as opposed to single seismometers, to determine whether the observed noise level might be decreased.

(C) A second area of research that has not been given proper consideration is a study of particle velocity signals induced in the ice from a distant, submerged CW acoustic transducer. However, it is our understanding that the General Motors Research Defense Laboratories will experiment soon in the Arctic with a low-frequency (40 cps) CW transducer. In connection with this experiment we recommend a study of the following problems: (a) the polarization of the orbital motion of the signal and the signal intensity as a function of range, which could provide a comparison with theoretical calculations, and (b) the possible effect of ice thickness on signal intensity.

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Two environmental aspects seriously affect the value of seismometers on Arctic ice for detecting and locating an underwater acoustic source. First, within the 10 to 100 cps bandwidth, dynamic variations of 30 to 40 dB in the ice particle velocity (noise) have been observed. Estimates of signal amplitude have not yet been attempted, but these noise level variations suggest that, at times, unfavorable signal to noise ratios will be experienced. Second, as a result of the Arctic Ocean's velocity gradient, acoustical signals propagate dispersively, and this effect complicates the problem of locating the source.

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